The circulation of the eastern tropical Pacific: A review

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Abstract

During the 1950s and 1960s, an extensive field study and interpretive effort was made by researchers, primarily at the Scripps Institution of Oceanography, to sample and understand the physical oceanography of the eastern tropical Pacific. That work was inspired by the valuable fisheries of the region, the recent discovery of the equatorial undercurrent, and the growing realization of the importance of the El Niño phenomenon. Here we review what was learned in that effort, and integrate those findings with work published since then as well as additional diagnoses based on modern data sets.

Unlike the central Pacific, where the winds are nearly zonal and the ocean properties and circulation are nearly independent of longitude, the eastern tropical Pacific is distinguished by wind forcing that is strongly influenced by the topography of the American continent. Its circulation is characterized by short zonal scales, permanent eddies and significant vertical transport off the equator. Notably, the Costa Rica Dome and a thermocline bowl to its northwest are due to winds blowing through gaps in the Central American cordillera, which imprint their signatures on the ocean through linear Sverdrup dynamics. The strong annual modulation of the gap winds and the meridional oscillation of the Intertropical Convergence Zone generates a Rossby wave, superimposed on the direct forcing, that results in a southwestward-propagating annual thermocline signal accounting for major features of the observed thermocline depth variations, including that of the Costa Rica Dome, the Tehuantepec bowl, and the ridge-trough system of the North Equatorial Countercurrent (NECC). Interannual variability of sea surface temperature (SST) and altimetric sea surface height signals suggests that the strengthening of the NECC observed in the central Pacific during El Niño events continues all the way to the coast, warming SST (by zonal advection)

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in a wider meridional band than the equatorially-trapped thermocline anomalies, and pumping equatorial water poleward along the coast.

The South Equatorial Current originates as a combination of equatorial upwelling, mixing and advection from the NECC, and Peru coastal upwelling, but the proportions and their variability remain unresolved. Similarly, while much of the Equatorial Undercurrent flows southeast into the Peru Undercurrent, a quantitative assessment is lacking. We are still unable to put together the eastern interconnections among the long zonal currents of the central Pacific.

This paper is part of a comprehensive review of the oceanography of the eastern tropical Pacific.

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1 Introduction

Spurred by the valuable tuna fishery of the region (Sosa-Nishizaki, Lluch-Cota, Dreyfus, Galvan, Ortega, and Castillo 2003 [this volume]) a major effort was undertaken by the Scripps Institution of Oceanography in collaboration with national and international fisheries research organizations in the late 1950s and early 1960s to observe the eastern tropical Pacific. Indeed, the opportunity presented by this program brought Klaus Wyrtki to Scripps in 1961 (Von Storch, Sundermann, and Magaard, 1999). The seminal papers describing the circulation, dynamics and water properties of the region came out of this project (Reid, 1948; Cromwell, 1958; Wooster and Cromwell, 1958; Cromwell and Bennett, 1959; Roden, 1961, 1962; Bennett, 1963; Wyrtki, 1964, 1965b, 1966, 1967; Tsuchiya, 1975). Following this productive period, relatively few *in situ* physical observations have been made: for example more than 30% of the total modern database of CTD profiles in the region east of 130°W between 30°S-30°N were taken before 1975. The literature shows a similar distribution. Until the recent interest in the eddies off Central America (see Willett and Leben 2003 [this volume]), there was a significant fall-off in publications discussing the circulation of the region after the mid-1960s; except for an ongoing effort by researchers at CICESE (Baumgartner and Christensen, 1985; Badan-Dangon, Robles, and Garcia, 1989) and studies of equatorial dynamics based on the TAO (Tropical Atmosphere Ocean; see Hayes, Mangum, Picaut,

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Sumi, and Takeuchi 1991; McPhaden and Co-Authors 1998) moorings and cruises along 110°W (e.g., McPhaden and Hayes 1990), only scattered papers were published on the low-frequency regional circulation between 1967 and 2002. Since the Wyrtki papers of the 1960s are still widely cited as the authoritative description of the regional circulation (e.g., Badan-Dangon 1998), it seems useful to review what was known at that time and what has been learned since to confirm, supplement, update or contradict that description. The circulation reviewed here is related to the physical forcing and water mass characteristics described in reviews of the east Pacific atmosphere (Amador, Alfaro, Lizano, and Magaña 2003 [this volume]), the variability of the surface heat fluxes (Wang and Fiedler 2003 [this volume]), the hydrography of the eastern Pacific (Fiedler and Talley 2003 [this volume]), the regional signatures of interdecadal variations and climate change (Mestas-Nuñez and Miller 2003 [this volume]), and the mesoscale eddies (Willett and Leben 2003 [this volume]).

Much of the descriptive and dynamical study of the tropical Pacific from the 1970s to the 1990s focused on the long zonal scales of the central Pacific, where the winds, thermal structure and currents are nearly a function of latitude alone, and where a zonal-average meridional section (with $\partial/\partial x$ taken to be zero) is a useful approximation. By contrast, in the east the presence of the American continent is the dominant factor, producing strong meridional winds and zonal variations of forcing and properties. In addition the necessity for the large zonal transports to be redistibuted as the currents feel the coast produces a complicated (and three-dimensional) structure. The focus of this review is on the circulation in the region where the shape of the continent and the continentally influenced winds are the dominant factor; for recent discussions of circulation in the central Pacific see McPhaden and Co-Authors (1998); Lagerloef, Mitchum, Lukas, and Niiler (1999); Johnson and McPhaden (1999); Johnson, Sloyan, Kessler, and McTaggart (2002). The region affected by the continent extends as much as 2500 km west of the coast of Central America, though much less than this north of 20°N or south of the equator. A physically-based way to define the region is where the southeast corner of the mean North Pacific Gyre does not reach into the large bight between central Mexico and Ecuador, thus the boundary stretches south from the tip of Baja California approximately along 110°W. This conforms to the southeasternmost edge of the gyre-like winds of the North Pacific and of the zonally-oriented Intertropical Convergence Zone (ITCZ) that is typical of the central Pacific (Fig. 1; see section 4.1). The wind forcing associated with the ITCZ between the two subtropical highs produces the long ridges and troughs that bound the tropical zonal currents of the central Pacific (South Equatorial Current (SEC), North Equatorial Countercurrent (NECC), and North Equatorial Current (NEC)). East of 110°W this ridge-trough system breaks down (Fig. 2, and see Fiedler and Talley 2003 [this volume]), and meridional flows and strong zonal gradients of both winds and currents are prevalent, with southerly cross-equatorial winds and a complex system of currents that exchange mass meridionally. The region between the tip of Baja California and the equatorial cold tongue also roughly coincides with the east Pacific warm pool (Wang and Fiedler 2003 [this volume]), so it makes sense to discuss it as a distinct dynamical and climatic regime. To the south, equatorial dynamics tie the entire equatorial region firmly together, so the eastern equatorial Pacific's circulation cannot be considered in isolation. As Wyrtki (1966) noted, physical features like the cold tongue extend far out into the central Pacific and do not have a well-defined boundary on their west. In addition, Kelvin waves efficiently carry the signatures of west-central Pacific wind anomalies to the east, and determine a large fraction of east Pacific thermocline depth variations (e.g., Spillane, Enfield, and Allen 1987; Kessler and McPhaden 1995b). However, winds in the far eastern equatorial Pacific are distinct in being characterized by a strong southerly cross-equatorial component, unlike the nearly zonal winds of the central basin (Fig. 1), which suggests a qualitatively different dynamical regime in the east. In keeping with the definition used for the region north of the equator, we will define the eastern part of the equatorial Pacific as the longitude range where the meridional component of the wind is larger than the zonal component; this boundary is approximately at 110°-105°W (Fig. 1). Much field study over the past two decades has been done in connection with the TAO mooring at 110°W, so this choice allows incorporation of that work in the review.

2 What Wyrtki knew: The picture in the mid-1960s

During the mid-1960s, Wyrtki assembled the physical picture that had been gained under the Scripps cruises and longterm ship-drift records into three partly-overlapping papers (Wyrtki, 1965b, 1966, 1967). These summary papers represented his own work, and also that of Cromwell (1958), Wooster and Cromwell (1958), Knauss (1960), Roden (1962), Bennett

(1963) and other Scripps collaborators. The historical shipdrift data (Cromwell and Bennett, 1959) tracked the surface currents and their seasonal variations in fair detail, and a series of expeditions conducted by Scripps, by government fisheries vessels, and under the CalCOFI program (see Table 1 of Wyrtki 1966) provided a few hundred bottle casts and a few thousand bathythermograph profiles in the region.

This by-modern-standards-limited dataset enabled Wyrtki to draw the schematic seasonal circulation shown in Fig. 3, which is in broadscale agreement with results from the picture that would be drawn from the much more comprehensive modern surface drifter data set (Fig. 4). Among the important features recognized were the winter extension of the California Current southward past Cabo San Lucas, the strengthening of the North Equatorial Countercurrent (NECC) during August-January, and its drastic weakening in the east during boreal spring (section 4.2), the mean Costa Rica Dome and associated Costa Rica Coastal Current (CRCC) (section 4.1), and the cyclonic circulation in the Panama Bight. Roden (1962) had shown the relevance of wind-driven Sverdrup dynamics to the eastern Pacific, especially in realizing that the Peru Undercurrent must be rather strong (making the total transport southward along the Peru coast), though he was limited by the poor quality of wind datasets available. Important missing elements in the mid-1960s picture were the weakness of the South Equatorial Current (SEC) in a narrow band right at the equator (especially during March-July) (section 4.3) and the connection between the NECC and the SEC (which is still poorly known). Although the Peru upwelling was well known and equatorial upwelling was surmised (Cromwell, 1953), the three-dimensional structure of the eastern Pacific was not well understood. Wyrtki was aware of the sharp SST front bordering the north edge of the equatorial cold tongue (Cromwell, 1953), but the tropical instability waves (TIW) that distort the front on monthly timescales (Baturin and Niiler, 1997; Chelton, Wentz, Gentemann, de Szoeke, and Schlax, 2000c) were not discovered until the advent of satellite SST in the late 1970s (Legeckis, 1977). Similarly, the strong eddies that occur under the winter winds blowing through the gaps in the Central American cordillera (Willett and Leben 2003 [this volume]) were not described until the 1980s, although the effects of these winds on the shape of the local thermocline were described by Roden (1961). The Costa Rica Dome was an early focus because of the importance of the tuna catch there, and Wyrtki (1964) devoted an entire paper to it. However, the lack of appreciation at that time for the role of wind-driven vorticity dynamics led him to drastically underestimate the upwelling into the Dome and to misunderstand its mechanism (see section 4.1).

A key factor that enabled a description of the circulation was the then-recent discovery of the equatorial undercurrent (EUC; Cromwell, Montgomery, and Stroup 1954, also see McPhaden 1986 and Montgomery 1959 for a history). By the time Wyrtki wrote his summary papers in the early 1960s, several expeditions had confirmed and added detail to Cromwell's discovery (Knauss, 1960), and it was obvious that the large eastward transport of the EUC must play a major role in the mass balance of the eastern Pacific. Previously, it had been hard to account for the westward growth of the SEC, which is much larger than the Peru Current that apparently fed it. Knauss (1960) estimated the EUC transport at 39 Sv, and Wyrtki (1966) used that value to draw a schematic circulation (reproduced here as Fig. 5), in which about 20 Sv of EUC water went into the SEC near the Galapagos. However, though realizing that substantial EUC water must feed the SEC, Wyrtki thought that a large upwelling transport was impossible, commenting "the amount of water upwelled along the equator certainly does not exceed a few 10^{12} cm^3/sec [Sv], otherwise the thermal structure would break down" (Wyrtki, 1966). Therefore, he postulated that the EUC water must move into the lower levels of the SEC without rising to the surface. Wyrtki corrected himself 15 years later with a since-confirmed estimate of 50 Sv of upwelling over 170°W-100°W, based on the meridional divergence of Ekman and geostrophic transports (Wyrtki, 1981). A remaining item of confusion that was not cleared up until much later concerned the spreading of isotherms about the EUC (Fig. 6 and section 4). It was thought at the time that this spreading must be a signature of vertical mixing, which seemed reasonable in such a strong current. However, turbulence measurements beginning with Gregg (1976) showed that mixing is in fact very weak at the EUC core; instead being strong in the stratified region above the current where it serves to maintain the balance between large upwelling transport and downward heat flux (see Gregg 1998 for a review of the turbulence observations, and Fiedler and Talley 2003 [this volume] for a description of the EUC hydrography).

3 Data and methods

Although much of this paper reviews published work, additional calculations are reported based on ocean and wind observations. These data sets and their processing are discussed in detail in the Appendix. The principal thermal data set used to construct the average annual cycle of thermocline depth (section 4.2) is a compilation of historical XBT profiles by Donoso, Harris, and Enfield (1994), and referred to here as the "AOML XBT data" (see Appendix, section A). Directly measured velocities are available from the ships servicing the TAO moorings and from a few of the moorings themselves (see Appendix, section C.1). The ship velocities are referred to here as the "Johnson ADCP data" (Johnson, Sloyan, Kessler, and McTaggart, 2002), and the moored velocities as the "TAO data". Satellite scatterometer winds are used to estimate Ekman pumping and to force a Rossby wave model (section 4.2.1). These winds were sampled by the European Research Satellite (ERS) over the period 1991-2001, used to construct an average annual cycle, and also by the Quikscat (Seawinds) instrument for the period August 1999 through July 2002 (see Appendix, section B) These are referred to here as the "ERS" or the "QuikSCAT" winds, respectively. Supplementary data sets (directly measured velocities, satellite SST, sea surface height (SSH) from satellite altimetry, and surface drifter velocities) are used to augment the principal data sources for particular purposes (see Appendix, section C).

4 Observations

The mean density structure and the resulting geostrophic circulation are shown in plan view and as meridional sections in Fig. 2 and Fig. 6, respectively (see also Fiedler and Talley 2003 [this volume]). (Dynamic height, shown in Fig. 2, measures the vertically-integrated density anomaly, expressing the fact that a lighter (i.e., warmer or fresher) water column of a given mass stands taller than a denser one. Because deep currents are observed to be small, it is a reasonable to assume that there are no pressure gradients at some deep reference level, so the mass of water above that level must be the same everywhere, and thus the density of each column determines its height. Dynamic height is scaled to accurately represent sea surface height relative to a reference level. Since most of the density contrast between adjacent columns arises because of thermocline variations, a water column with deep thermocline has a thick upper warm layer and is overall warmer than a column with shallow thermocline; thus it has higher dynamic height. Dynamic height is a convenient quantity because its contours are streamlines of the geostrophic flow, and its cross-stream gradient measures the flow speed; these relations are seen in Fig. 2.) At the western edge of the region, the well-known currents of the central Pacific are seen. The thermal structure at 125°W (Fig. 6, top) shows the thermocline sloping up towards the equator in the southern hemisphere, giving a broad and strong westward geostrophic SEC, then isotherm spreading about a center at 80m which is the signature of the eastward EUC. The SEC spans the equator at the surface; although the upward-bowing isotherms above about 18°C (the expression of equatorial upwelling) qualitatively indicate a geostrophic westward current, frictional and nonlinear terms are also important in the balance (Philander and Pacanowski, 1980; Yu, Schopf, and McCreary, 1997), as is the mixing mentioned in section 2. The trough at 5°N and the ridge at 10°N define the eastward NECC, and the downward slope north of the ridge gives the westward NEC, which is the southern limb of the North Pacific subtropical gyre. Below the thermocline, the paired eastward currents at about 125-400m depth near ±4-5° latitude, associated with the deep bowl of isotherms from 10°-12°C (Fiedler and Talley 2003 [this volume]), are the Subsurface Countercurrents (Tsuchiya Jets) that originate in the far western Pacific (Tsuchiya, 1975; Rowe, Firing, and Johnson, 2000; McCreary, Lu, and Yu, 2002). All these currents can be seen entering and leaving the region at the western edge of Fig. 2. Fig. 6 (top) is similar to sections that would be made as far west as about 170°E, except that all the surface currents would be somewhat stronger further west (Wyrtki and Kilonsky 1984; Taft and Kessler 1991; Johnson, Sloyan, Kessler, and McTaggart 2002; Fiedler and Talley 2003 [this volume]).

4.1 Northeastern tropical Pacific: Mean circulation

East of 120°W the picture is much more complicated and includes substantial meridional flow that feeds and drains the outgoing and incoming zonal currents. The California Current flows south along the coast of Baja California and turns west between 12°-20°N into the NEC to mark the southeastern corner of the North Pacific subtropical gyre (Fig. 2); at thermocline level (bottom panel) the tip of the gyre is clear but at the surface (top panel) the geostrophic currents continue southeastward along the coast of Mexico. However, the Ekman transport in this region is westward (i.e., to the right of the wind vectors in Fig. 1) so the total surface flow is southwest (Fig. 4), approximately parallel to the geostrophic contours at 100m.

The Costa Rica Coastal Current is the strong (20 cm s^{-1}) northwestward current between the Costa Rica Dome and the coast that forms the poleward limb of the dome. The lack of ocean data along the west coast of Mexico has led to considerable confusion about the path of this current after it leaves the dome region. Wyrtki (1965a) showed the CRCC continuing all the way to the mouth of the Gulf of California during boreal fall (as redrawn in Fig. 3), and that work continues to be cited (e.g., Baumgartner and Christensen 1985; Badan-Dangon 1998). The somewhat more complete modern data sets studied here do not support that interpretation, instead suggesting that the mean CRCC feeds the NEC directly, at about 10°-12°N south of Tehuantepec, and mean flow along the Mexican coast from Jalisco to Oaxaca is in fact southeastward (Fig. 2, top, and Fig. 4). In section 4.2 below, we find that this southeastward current appears to prevail all year except during August-September, when the geostrophic flow is weak, and the CRCC does not extend north past Tehuantepec at any time during the climatological year (e.g., see Fig. 8 below). However, neither the hydrographic nor the drifter data sets are sufficiently well-sampled near the Mexican coast to make that conclusion unambiguous, and further investigation in this region will be necessary before these currents can be reliably charted.

In addition to the CRCC, the NEC appears to have other sources: the California Current, and flow that has turned north from the NECC at about 110°-120°W. Both Fig. 2, top, and Fig. 4 show that the source region of the NEC exhibits a complex interleaving of currents. In fact these average pictures represent a mix of seasonally varying influences (see section 4.2).

At 110°W, the EUC and SEC are similar to 125°W, though perhaps 20m shallower, but the mean geostrophic NECC is nearly absent (Fig. 2, top and Fig. 6, middle). The ridge-trough system that supports the NECC has almost completely flattened out (Fig. 6, middle), and the weak eastward geostrophic flow found there is due to isotherm slopes below 13°C; that is, associated with the Tsuchiya Jet and not the thermocline. The surface dynamic height shows virtually no meridional pressure gradient at 110°W between 4°N and 9°N, and the NECC appears to split and turn south into the SEC and north into the NEC (note the 78 and 80 dyn-cm contours in Fig. 2, top). This is a remarkable termination for a current well out in the interior basin, and is contradicted by the drifter tracks, which show an apparently-continuous NECC all the way from 120°W to the Costa Rica Dome (Fig. 4). The discrepancy is not due to sampling or other problems with the XBT data, as the directly-measured currents are nearly identical to the geostrophic currents (compare Fig. 6, middle, which uses measured currents south of 9°N, with Fig. 2, which is entirely geostrophic). Kessler (2002) noted that the Ekman flow under the region's southerly winds (Fig. 1) is strongly eastward, in part due to the small value of the Coriolis parameter this close to the equator. With reasonable assumptions about the Ekman depth (Ralph and Niiler, 1999), the Ekman currents account for nearly all the difference between the geostrophic (Fig. 2) and drifter (Fig. 4) depictions of the NECC at 110°W (see section 4b and Fig. 4 of Kessler 2002) Thus the apparently-continuous NECC seen by the drifters is a near-surface Ekman feature, driven by the southerly winds, that does not represent the geostrophic NECC that can be followed across the entire basin. Since the Ekman flow is quite shallow the transport of the eastward jet in Fig. 4 is small. On the other hand it will be shown in section 4.2 below that during boreal fall there is a geostrophic NECC east of 110°W, and that the mean situation seen in Fig. 2 is an average over a strongly-varying seasonal pattern.

A striking bowl and dome are found in the mean dynamic height, centered at 13°N, 105°W and 9°N, 90°W respectively (Fig. 2, top). (In the following, we will refer to these features as they appear in thermocline topography, not dynamic height; that is, the Costa Rica Dome at 9°N where the 20°C isotherm rises to about 25m, and the "Tehuantepec bowl" at 14°N where 20°C is at about 90m). Since the upwelling associated with the Costa Rica Dome produces a nutrient-rich environment that supports a large fishery (Sosa-Nishizaki, Lluch-Cota, Dreyfus, Galvan, Ortega, and Castillo 2003 [this volume]), it has been studied much more intensely than the Tehuantepec bowl, which does not even have a recognized name. These features appear to be the eastern ends of the thermocline trough and ridge that define the limits of the NEC across the basin, with the trough connecting to the center of the subtropical gyre and the ridge running along about 9°N between the NEC and NECC all the way to the Philippines (Wyrtki, 1975b; Kessler, 1990). Notably, however, both the trough and ridge flatten near 110°W (as has been commented on above with respect to the NECC) before restrengthening near the coast and terminating in the nearly detached bowl and dome. Both features are clearly seen as counter-rotating eddies in the drifter (Fig. 4) and geostrophic (Fig. 2, top) velocity vectors.

Wyrtki (1964) noted the cool SST, low oxygen and high salinity and phosphate values in the center of the Costa Rica Dome as an indication of the emergence of subthermocline water via upwelling. Based on sections from several cruises, he estimated the current speed around the dome at about 20 to 50 cm s^{-1}, which is comparable to modern estimates (e.g. Fig. 4). Wyrtki hypothesized that the centrifugal acceleration around the Costa Rica Dome would contribute an outward divergence and consequent upwelling. The upwelling estimated this way was quite small, less than 0.1 Sv, about 1/30th the modern value, although the values of all the terms Wyrtki used were similar to those that would be used today. Wyrtki's assumption that the reason for the upwelling was rapid rotation around the eddy as the NECC turned at the coast, was the source of the error. If he had scaled the terms he would have found that the centrifugal term is at least ten times smaller than the geostrophic term. In fact, Costa Rica Dome upwelling is driven by the wind, as discussed below, which Wyrtki (1964) dismissed as weak and variable in the absence of sufficient observations, and he omitted wind forcing from his balance.

The distinctive regional wind forcing is key to understanding the complex thermal structure of the region and the way the zonal currents of the central Pacific become modified near the continent. West of about 110°W the characteristic central Pacific winds are seen (Fig. 1), with trade winds converging into a well-developed ITCZ. East of 110°W and north of the equator the pattern is quite different: instead of a zonally-oriented ITCZ, the winds (and especially the wind curl) are dominated by jets blowing through three gaps in the Central American cordillera: the Chivela Pass at the isthmus of Tehuantepec in Mexico, the Lake District lowlands of Nicaragua inland of the Gulf of Papagayo, and the central isthmus of Panama where the Panama Canal was built (Fig. 1). The wind stress curl is $Curl(\tau) = \partial\tau^y/\partial x - \partial\tau^x/\partial y$, where the components of the wind stress vector τ are τ^x and τ^y; it expresses the rotation a particle would experience in a wind field that varies in space. The Central American wind jets extend at least 500 km into the Pacific and produce distinctive curl dipoles as the wind strength decreases away from the jet axis: each jet has a region of positive curl on its left flank and negative curl on its right. (In the mean, the wind jets are more clearly defined in the curl dipoles than in the vector winds themselves; Fig. 1). The magnitudes of these curls are at least as large as that of the ITCZ. The positive curl on the south flank of the Papagayo jet is enhanced and extended to the west because of the westerly winds south of the jet (Mitchell, Deser, and Wallace, 1989). The three wind jets are known to vary on short (weekly) timescales, especially in association with winter high pressure systems transiting North America (Chelton, Freilich, and Esbensen, 2000a,b), producing oceanic eddies of various types (Willett and Leben 2003 [this volume]). For present purposes, we are interested in the jets' impact on the low-frequency dynamics, in which Ekman pumping due to their curl is the main factor.

Ekman pumping occurs because the winds and therefore the Ekman transport vary spatially, which produces convergences and divergences in the upper layer. As a consequence, the thermocline must rise or fall to maintain mass balance, so Ekman pumping is interpreted as a vertical velocity at the base of the surface layer. Zonal and meridional Ekman transports are $(U_E = \tau^y/f\rho,\;\; V_E = -\tau^x/f\rho)$, where f is the Coriolis parameter and ρ the density. The divergence of the Ekman transports equals the wind stress curl divided by $f\rho$ (where f is for the moment taken as constant). The Ekman pumping velocity $Curl(\tau)/f\rho$ is of fundamental importance for the ocean circulation because it produces thermocline depth variations and resulting pressure gradients, which consequently produce geostrophic flow. If not for Ekman pumping of the thermocline, ocean currents would occur only as directly-driven frictional flows within the mixed layer. For example, under northern hemisphere wind jets like those west of Central America, the Ekman transport is to the right of the wind direction, and is largest at the jet axis. Approaching the jet from its left, the Ekman transport is increasing, so on the left flank of the jet it is divergent, leading to upwelling. To the right of the jet axis, the Ekman transport is decreasing, and is thus convergent (downwelling). The curl dipoles seen in Fig. 1 thereby produce the thermocline bowls and domes, and the corresponding highs and lows in dynamic height (Fig. 2, top).

The consequences of Ekman pumping go beyond changing the thermocline depth, however. On a rotating planet, a locally still water column has the rotation rate ("vorticity") about its vertical axis equal to the local value of the Coriolis parameter divided by two (thus it is zero at the equator and grows in magnitude towards the poles). If the column is lengthened or shortened by Ekman pumping, its vorticity will be changed by changing its cross-sectional area. Stretching produces increasing vorticity because the column becomes thinner and the same angular momentum is achieved by a faster rotation. In steady state (where vorticity is conserved) a local increase in vorticity is balanced by moving poleward, where the vertical component of the earth's rotation (the planetary vorticity) is larger. This is the basis of the Sverdrup relation, $\beta V = curl(\tau)$, where $\beta = df/dy$ is the meridional gradient of the Coriolis parameter and V is the total meridional (Sverdrup) transport (Sverdrup, 1947). For example, under positive curl, as occurs on the south flank of the Papagayo jet, Ekman pumping lifts the thermocline and thus stretches the water column beneath. The vorticity of the column increases, and it must move poleward to a latitude where its spin equals the planetary vorticity. Although the Sverdrup relation applies to the vertical integral, if the vertical structure of the geostrophic flow is independently known (in this case from the XBT data), then the vertical structure of vertical motion induced by Ekman pumping can be deduced.

The region of positive wind curl on the south flank of the Papagayo jet (Fig. 1) produces strong upwelling (10-20 m month^{-1}, comparable to equatorial upwelling) because the Ekman transport is northward under the easterly jet itself at 10°-11°N and southward under the westerly winds at 6°-8°N. As a result of this surface divergence, the thermocline is lifted and the water column beneath is stretched, forming the Costa Rica Dome at 9°N (Fig. 7, top, shows a zonal slice and Fig. 6, bottom, a meridional slice through the center of the dome). The thermocline dome produces a cyclonic eddy-like geostrophic circulation at the surface (Figs. 4 and 2 (top)), but no dome is seen below the shallow thermocline. Instead, the subthermocline isotherms slope up to the west over a broad region to at least 105°W (Fig. 7, top, and see Fiedler and Talley 2003 [this volume]), and the resulting geostrophic flow deeper than about 50m under the dome is all northward (Fig. 7, bottom, or Fig. 2, bottom). The vertically-integrated transport is northward, quantitatively consistent with the Sverdrup relation under positive curl (Kessler, 2002), and the cyclonic eddy circulation is a very shallow feature. Knowing the vertical structure of geostrophic transport (from XBT data) allows the structure of vertical motion to be inferred (if both v and w are assumed to be zero at 450m); this is found to be a transport of about 3.5 Sv upward across the 17°C isotherm into the dome, where it diverges in the Ekman outflow (Kessler, 2002). Thus the dome and the northward flow beneath it are both part of the ocean response to the Papagayo jet. The dome is one of the few places in the ocean where upwelling occurs mostly from below the thermocline (equatorial upwelling occurs primarily from the upper levels of the thermocline), and thus constitutes a perhaps-important means of communication from the intermediate to the surface layer. The isotherms would continue to be advected upward, if not for surface mixing under the strong wind jet, which maintains the steady balance. Wyrtki (1964) did not realize the importance of these vorticity dynamics because the existence of the Papagayo wind jet and its curl were not known at the time, and because the very sparse thermal observations available concentrated on the position of the surface dome itself and did not extend west of about 91°W, so the clue of northward transport under the dome remained unknown.

The Costa Rica Dome upwelling and subsurface northward transport are also linked with the northern Tsuchiya Jet that flows across the entire Pacific at about 4°N, 150-200m depth (Fig. 6). The northward flow under and to the west of the dome is the Tsuchiya Jet turning abruptly away from the equator (Fig. 2, bottom, and note the disappearance of both Tsuchiya Jets at 90°W, compared to the sections further west in Fig. 6). About half the roughly 7 Sv transport of the northern Tsuchiya Jet upwells into the Costa Rica Dome, while the other half turns and flows west into the lower reaches of the NEC (Fig. 2, bottom; note the large Tsuchiya Jet approaching the dome and weaker flow leaving the region at these depths). The mechanism driving the Tsuchiya Jets remains in question, but recent theory suggests that the northern branch is in fact driven by Costa Rica Dome upwelling (McCreary, Lu, and Yu, 2002), as an example of a "β-plume" (Stommel, 1982). The southern branch may be similarly driven by upward motion across a broader region of the southeast Pacific (the negative curl along 5°S in Fig. 1), with a subthermocline dome near 10°S, 85°W (Voituriez, 1981). The Atlantic also has Tsuchiya Jets flowing into upwelling in the Guinea and Angola Domes (Mazeika, 1967). Thus Costa Rica Dome upwelling may be much more than a local feature of the eastern Pacific.

The mean ocean response to the Tehuantepec and Panama wind jets is also consistent with the Sverdrup relation: northward geostrophic transport is found under their left flanks and southward under their right. (Unlike Papagayo, for these jets the downwelling curl is larger than the upwelling curl, Fig. 1). Note the northward bulge of the 76 dyn-cm contour into the Gulf of Tehuantepec (Fig. 2, top), and the cyclonic circulation in the Gulf of Panama (most evident at 100m in Fig. 2, bottom) (Stevenson, 1970). The Tehuantepec bowl at 13°N, 105°W is seen as a large, permanent anticyclonic eddy both in the geostrophic circulation (Fig. 2) and the drifters (Fig. 4). The bowl is consistent with a linear response to the downwelling curl on the right flank of the Tehuantepec wind jet (Kessler, 2002). Interpreting the mean and low-frequency evolution of the Tehuantepec bowl from sparse XBT data is tricky because it is in an area where anti-cyclonic eddies pass each winter (Willett and Leben 2003 [this volume]), and it might be that the mean bowl was simply a reflection of sampling many such eddies. In addition, the strength of the short-timescale eddies suggests that nonlinear terms would be important, however this seems not to be dominant for the mean circulation, which can be reasonably well diagnosed based on linear Sverdrup dynamics (Kessler 2002, and see section 4.2). In the Gulf of Panama, cool SST along the coast of Colombia is a signature of upwelling under the positive curl.

4.2 The annual cycle in the northeastern region

Although the mean situation in the eastern tropical Pacific is quite different from the central basin, the annual cycle variations are similar in phase and amplitude to those much further west (Kessler, 1990). Annual cycle thermocline anomalies consist of an out-of-phase relation across a line slanting from 8°N, 120°W to 3°N, 90°W, a continuation of a nodal line that extends along roughly 8°N to about 170°W (Kessler 1990, and see Fig. 7b of Fiedler and Talley 2003 [this volume]). These thermocline variations appear entirely consistent with those of the wind stress curl, which also varies out of phase across 8°N from the Philippines east to 90°W, as the upwelling curl of the ITCZ moves north and south with the sun. In November, the thermocline is shallow north of the line and deep south of it, following several months in which the ITCZ is at its most northerly, and the reverse occurs in May. Since this nodal line runs roughly along the axis of the NECC, the result is to increase the thermocline slope across the NECC in November, and weaken it in May (Fig. 8). To a lesser degree, these anomalies also strengthen and weaken the NEC and SEC at the same time. During November, the NECC is strong across the basin; it flows eastward directly around the Costa Rica Dome and then into the NEC (Fig. 8, lower right). Both these currents are entirely zonally oriented at this time, and appear to be a nearly closed system. During May, the situation is quite different. The geostrophic NECC is absent (flow in this latitude range is actually reversed) from 130°W to 100°W, and the NEC is fed instead by flow coming south from the California Current, and clockwise around a much-strengthened Tehuantepec bowl, which has moved somewhat further offshore at this time of year (Fig. 8, lower left, and see the results of the Rossby wave model in section 4.2.1). Much more southward geostrophic flow is observed in the first half of the year, apparently allowing water from the California Current to penetrate far to the south. Water property or tracer analysis might be able to determine if boreal spring is a window that allows communication from the mid-latitude to the tropical eastern Pacific, as it appears from the varying currents.

Fiedler (2002) used climatological temperatures and winds to diagnose the annual cycle of the Costa Rica Dome, similar to the sequence seen in dynamic heights found from the AOML XBT data (Fig. 8). Fiedler's data showed that thermocline uplift begins at the coast in February-April as strong upwelling wind stress curl occurs on the south flank of the Papagayo wind jet. In May, the Papagayo winds weaken and the dome separates from the coast. During July-October, as the ITCZ moves north, upwelling curl lifts the 10°N thermocline ridge across the entire basin, including the eastern tropical Pacific, and the ridge strengthens and becomes more closely connected with the Costa Rica Dome, which appears to have lengthened to the west. During November-January, the ITCZ moves south, and northeast trade winds blow strongly over the region; the dome deepens to its weakest values by January. However, the core region of the dome, at 90°W, has relatively little annual variation, with most of the changes seen in its westward expansion and contraction (Fig. 8), and the CRCC remains strong throughout the year. This lack of variability occurs because upwelling curl within about 200km of the coast is provided by the Papagayo jet on the north side of the dome in winter and by ITCZ westerlies on the south side of the dome in summer. Further west the curl alternates during the year with the ITCZ migration. The Tehuantepec wind jet is strongest in boreal winter, which produces upwelling curl to the north of the dome; this upwelling shoals the northern edge of the dome while its center is deepening (note the northward bump on the dome in the lower right panel of Fig. 8).

Fiedler's linear, wind-driven interpretation of the Costa Rica Dome annual cycle agrees with diagnoses made from much cruder data by Hofmann and O'Brien (1981), but others have suggested that nonlinearities are also important. Umatani and Yamagata (1991) argued that cyclonic eddies produced near the coast by strong Papagayo winds "seed" the growing Costa Rica Dome and are an essential element of its formation. In the next section, we use a simple model consisting only of linear long Rossby waves to suggest that although the eastern tropical Pacific is rich with seasonal eddies due to the wind jets, the low-frequency dynamics evolves principally according to a linear interpretation.

4.2.1 A Rossby wave model

With much more complete data coverage, it would in theory be possible to diagnose the terms of the equations of motion directly and thereby determine the importance of nonlinearity in the evolution of the annual cycle. Lacking such data, we take the tack of studying solutions to simple models to evaluate their consistency with observations. If a simple model is able to reproduce the observed phenomena, then we can conclude that, to the accuracy of the data, there is no justification for invoking more complex hypotheses. In situations where the simple model fails, it points to locations where other processes are active. The simplest first guess at the low-frequency, large-scale evolution of the off-equatorial thermocline is a model consisting only of long quasi-geostrophic Rossby waves forced by wind stress curl. This model adds elementary time dependence to the Sverdrup dynamics discussed in section 4.1, allowing time-varying wind stress curl to pump the thermocline depth. By its neglect of velocity acceleration terms, the model excludes the equatorial waves that would be essential to study the region less than about 3°-4° latitude, but it has proven useful in the tropics. The model can be written:

$$\frac{\partial h}{\partial t} + c_r \frac{\partial h}{\partial x} + Rh = -Curl\left(\frac{\tau}{f\rho}\right) \tag{1}$$

where h is the thermocline depth (positive down), τ is the wind stress and ρ the density of seawater. The long Rossby wave speed is $c_r = -\beta c^2/f^2$ (c is the internal long gravity wave speed, f is the Coriolis parameter and β its meridional derivative), and R is a damping timescale. Note that f is now allowed to vary with latitude. The two parameters to be chosen are the gravity wave speed c, which is estimated to be between 2 and 2.5 m s^{-1} in the tropical Pacific (Chelton, deSzoeke, Schlax, ElNagger, and Siwertz, 1998), and the damping timescale R, typically taken to be (6 to 12 months)$^{-1}$ (Picaut, Menkes, Boulanger, and duPenhoat, 1993). Here we choose c=2 m s^{-1} and R=(9 months)$^{-1}$; in fact the results are qualitatively insensitive to these choices within the reasonable ranges 1.75 m s^{-1} $\leq c \leq$3 m s^{-1} and $R \leq$ (2 years)$^{-1}$. Additional realism could perhaps be achieved by choosing a different gravity speed c at each latitude, or by letting c be a function of longitude as well, but this did not seem necessary for the present purposes of making a first guess at the importance of the linear response to wind forcing.

Since long Rossby waves propagate nondispersively due west, the wind-driven solution can be written separately at each latitude as an integral in x that sums the contributions of the wind forcing on the wave as it travels westward at speed c_r:

$$h_I(x_0,t) = -\frac{1}{c_r}\int_{x_E}^{x_0} e^{\frac{R}{c_r}(x-x_0)} Curl\left(\frac{\tau(x, t+\frac{x-x_0}{c_r})}{f\rho}\right) dx \tag{2}$$

where $h_I(x_0,t)$ is the interior wind-driven part of h at positions and times (x_0,t). Note that $Curl(\tau/f\rho)$ in (2) is evaluated not at time t but at previous times looking back along the wave ray at speed c_r; that is, at times $t-(x-x_0)/c_r$. The lower limit of integration x_E is the longitude of the eastern boundary.

Rossby waves can also radiate from the eastern boundary (for instance due to reflection of equatorial Kelvin waves), and these influences must be added to the interior solution (2):

$$h_B(x_0,t) = e^{\frac{R}{c_r}(x_E-x)} h_E\left(t+\frac{x_E-x}{c_r}\right) \tag{3}$$

where h_B is the damped effect of eastern boundary signals propagating to the interior points (x_0,t). As in (2), the value of h at the eastern boundary (h_E) is evaluated at previous times reflecting the lag for propagation from x_E to x_0. The complete solution h to (1) is h_I from (2) plus h_B from (3); these are solved at each latitude independently and then combined. Here we use the average annual cycle of the ERS winds (see Appendix, section B) to force (2), and the eastern boundary value of observed 20°C depth from the XBT data as h_E in (3). (In fact the boundary contribution to the total is small, except very close to the coast). Solutions to (1) have also been found using other scatterometer winds (Quikscat) and from *in situ* wind products (FSU) for various time periods, and the results are not strongly dependent on the wind data set used.

The results of the Rossby model are compared to observed 20°C depth anomalies in Fig. 9, for four average seasons. The solution can be thought of as a combination of westward-propagating Rossby waves forced by wind stress curl that seesaws annually across a nodal line at 8°N. The result is a generally southwestward phase propagation with wave crests approximately parallel to the Central American coast. In both the Rossby model and the observations, signals appear to originate as a thin region near the coast, grow and separate from the coast, and finally leave the region at the southwest corner. In general the model represents the magnitude and position of the observed thermocline signals reasonably well, though there is a sense that the model propagation is too slow at the northern edge and too fast at the southern (e.g., Kessler 1990). It might be possible to improve this by varying the gravity wave speed c, but it seems useful to appreciate the correspondence even with the simplest type of model.

A difference between model and observations is that the model solution depicts particular latitudes as having strong maxima, while the observations are smoother (Fig. 9). Since the model solutions are entirely independent at each latitude, its results are sensitive to narrow areas of strong wind stress curl, especially those associated with the mountain-gap wind jets (note the three positive maxima extending westward from the Tehuantepec, Papagayo and Panama jet outflow regions in the model JFM season). In reality, energetic eddies generated by the wind jets (but not represented in the linear model) produce mixing that blends latitudes together.

For the case of the Costa Rica Dome, the model correctly depicts the sequence of uplifting beginning at the coast early in the year, westward growth and separation in April-May-June, and strengthening of the ridge to the west of the dome in boreal summer-fall, as described by Fiedler (2002). The shoaling on the north side of the dome as its center weakens in boreal winter is also evident in the linear model (lower right panel of Fig. 9). For the case of the NECC, the model shows that observed deepening along the 10°N ridge near 110°W in April-May-June that weakens the NECC (Fig. 8, lower left) is consistent with the wind forcing and Rossby wave propagation (Fig. 9, second row). In boreal fall the wind-forced Rossby wave produce the opposite anomalies (Fig. 9, bottom row), and the NECC strengthens (Fig. 8, lower right).

There is little indication of the features of the mean thermocline topography in the observed annual cycle anomalies (Fig. 9, left panels). No signatures of the mean ridges and troughs can be seen, nor of the Costa Rica Dome, nor of the regions of strong eddy activity. Instead the observations show a smooth southwestward propagation right across the strong thermocline topography and current shears. The good agreement with the model solution, which assumes a flat background thermocline (by the choice of a single value for c), with no eddy mixing, is evidence for the major role played by the linear response to wind forcing in the evolution of the annual cycle.

The thermocline bowl centered near 14°N, 105°W (termed "Tehuantepec bowl" here) has a larger annual cycle amplitude than the Costa Rica Dome (Fig. 8), but has not been as well described in the literature. The bowl is nearly absent in boreal summer-fall, and grows as an isolated feature during boreal winter, with the 20°C isotherm at least 10m deeper than its surroundings. During boreal spring, the thermocline trough at 15°N amplifies and appears to extend eastward as the bowl moves west, connecting the two. In the sequence of anomalies (left panels of Fig. 9), this can be seen as the deep thermocline centered at 13°N, 100°W in the JFM season that extends as a long trough to the west in April-May-June, then shoals strongly in the JAS season. It might well be thought that this feature was a reflection of the large anticyclonic Tehuantepec eddies that are due to the winter wind jet (Willett and Leben 2003 [this volume]). It would certainly be possible that the deep thermocline seen in the XBT data was simply an aliasing due to sparse sampling of many warm-core eddies in boreal winter. In that case one would expect that the linear Rossby wave model would not represent this feature very well. However, the model solution in fact captures this aspect of the seasonal evolution quite realistically, suggesting that the boreal spring deepening is due to downwelling curl associated linearly with the western side of the Tehuantepec wind jet, and the boreal summer shoaling is due to the upwelling curl from the eastern side of the jet that arrives later (Fig. 9, top and upper middle set of panels). The downwelling curl from Tehuantepec extends over the bowl (e.g. Fig. 1), so this is felt during winter and soon after. The Rossby wave speed is about 190 km month^{-1} at 14°N, so the upwelling signal takes about 6 months to propagate from the east side of Tehuantepec to 105°W, weakening the bowl in summer. Further, the connection to the 15°N thermocline trough that is established in boreal spring (Fig. 8, lower left) is associated with the westward passage of the downwelling Rossby wave past the bowl (Fig. 9, upper middle panels). Though the region just west of the Gulf of Tehuantepec that is strongly affected by the winter eddies might be expected to be a place where linear dynamics are least appropriate, that turns out not to be the case.

We can conclude from the Rossby model comparisons that these simple dynamics gives a good first-order picture of the annual cycle evolution in the northeastern tropical Pacific. The large-scale features of the regional evolution are principally explained by a linear ocean response to wind forcing, despite the evident possibilities of a fundamental role for the energetic eddies and more complex dynamics. That means, for example, that one should not see the annual cycle of the Costa Rica Dome as an isolated phenomenon. Instead, the westward growth of the Dome and its separation from the coast reflect the passage of the regional-scale Rossby wave seen in Fig. 9, driven by the regional-scale winds. Discrepancies are found in the model wave being broken into specific latitude bands while the observations are smoothly connected across latitudes; presumably this would be improved by a model that allowed communication across latitudes by mixing, as occurs in the ocean.

4.3 Eastern equatorial Pacific: Mean and annual cycle

Relatively much less is known about the circulation in the far eastern than in the central equatorial Pacific, and a full picture of the currents and their interconnections is yet to be accomplished.

At the surface, drifter tracks (Fig. 4 and Reverdin, Frankignoul, Kesternare, and McPhaden 1994) suggest that the SEC is weak east of about 85°W, and rapidly gains speed as it flows west. This was described by Wyrtki (1966), who correctly interpreted the increase as a sign that the SEC was gaining mass from the decelerating EUC below (as noted in section 2 the role of the vertical circulation was not understood until later). Although there are no data to allow an estimate of the transport of the SEC east of 95°W, where it appears to be quite small (Figs. 6, bottom, and 4), the Johnson ADCP data (Appendix, section C.1) allows estimation of SEC transport on the 15° longitude spacing of the TAO lines, beginning at 95°W. SEC transport above 200m within 6°S-6°N at 95°W is about 12 Sv, increasing to about 35 Sv by 140°W, consistent with this interpretation.

Both the drifters (Fig. 4) and the ADCP data (Fig. 6) show that the SEC is split into two lobes, and is thin and weak on the equator, where the EUC surfaces or nearly so. Just south of the equator from the Galapagos to about 110°W the mean surface zonal flow is near zero, and a strong southward divergence from the equator is seen (Fig. 4). Although the drifter population is smaller in this region (because the drifters diverge rapidly from the equator), sampling problems do not appear to explain this unexpected feature. The same weak surface current appears in the Johnson, Sloyan, Kessler, and McTaggart (2002) ADCP data and is consistent with the very small mean surface zonal currents measured by the TAO mooring at 0°, 110°W (Kessler, Rothstein, and Chen, 1998). The splitting of the SEC has been attributed to upward advection of eastward momentum from the EUC, and interpreted as an important nonlinear effect of easterly equatorial winds (Philander and Pacanowski, 1980; McPhaden, 1981), but the weaker easterlies over the cold tongue, caused by atmospheric boundary layer effects of the cold SST (Chelton and Co-Authors, 2001), may also contribute. Another factor is the seasonal cycle. Although the zonal winds remain easterly throughout the year, they weaken by a factor of about two in March at 110°W (Yu and McPhaden, 1999a). The EUC surfaces during boreal spring, changing the near-surface flow at 0°, 110°W to be eastward at more than 20 cm s^{-1} during the average April and May (Kessler, Rothstein, and Chen, 1998; Yu and McPhaden, 1999a,b); this reversal reduces the annual mean zonal current to near zero. The annual shoaling of the EUC is a basinwide phenomenon that begins in the far east in March, and reaches the dateline by about August (Reverdin, Frankignoul, Kesternare, and McPhaden, 1994; Yu and McPhaden, 1999b; Johnson, Sloyan, Kessler, and McTaggart, 2002); it is forced by the westward-propagating annual cycle of zonal wind along the equator producing a mixture of equatorial Kelvin and long Rossby wave modes (Yu and McPhaden, 1999b). Wind-forced ocean models show this signal clearly (e.g., Yu, Schopf, and McCreary 1997 and Kessler, Rothstein, and Chen 1998). Because the mean EUC is much shallower in the east, it is only here that an actual reversal of the surface westward flow is seen. During boreal spring, surface flow along the equator is eastward from about 140°W to at least 95°W (Johnson, Sloyan, Kessler, and McTaggart, 2002). It is worth noting that the seasonal cycle has a low-frequency modulation and has been observed to change substantially in amplitude in different decades (Gu, Philander, and McPhaden, 1998).

The fact that the weakest SEC is found south of the equator (Fig. 4 and Fig. 6, bottom) is consistent with the meridional circulation induced by southerly winds (Fig. 1); note that the coldest SST is also south of the equator (Fiedler and Talley 2003 [this volume]). These cross-equatorial winds drive a cell that is downwind (northward) at the surface and upwind in the thermocline, with enhanced upwelling south of the equator (Philander and Delecluse, 1983; Mitchell and Wallace, 1992; Kessler, Rothstein, and Chen, 1998). This cell therefore advects westward wind-input momentum northward across the equator, and upwells eastward momentum from the EUC at about 1°-2°S, contributing to the asymmetric pattern seen in Fig. 4.

A comprehensive examination of available hydrographic data by Lukas (1986) showed that high-salinity, high-oxygen water flowing eastward in the EUC is advected to the coast of Ecuador and Peru (Fiedler and Talley 2003 [this volume]), indicating that the EUC continues further east than the Galapagos. The implication is that this water feeds the Peru Undercurrent and becomes a source of the Peru upwelling. Toggweiler, Dixon, and Broecker (1991) confirmed Lukas' result using ΔC^{14} measurements from coral heads. They suggested that low ΔC^{14} water originates in the subantarctic region of the southwest Pacific, flows northward to enter the lower reaches of the EUC, and populates the east Pacific thermostad to eventually upwell off Peru. Steger, Collins, and Chu (1998) found evidence of the EUC splitting around the Galapagos with the main branch flowing to its south. The region was crossed at 86°W by WOCE cruise P19 in March 1993 (Tsuchiya and Talley, 1998) which ran an underway ADCP profiler; this showed what appears to be a strong EUC (70 cm s^{-1}) centered just north of the equator at 70m depth. However, this period was during a weak El Niño that produced several Kelvin wave pulses of eastward equatorial currents that were observed further west around this time (Kessler and McPhaden, 1995b), so the single P19 section may very well not represent the mean. It should be suspected that the EUC would be found south of the equator because of the meridional circulation cell mentioned above in connection with the SEC. Although at thermocline depth the axis of the EUC appears to turn southeast towards Peru, there may be substantial export to the northern hemisphere due to strong lateral mixing by the tropical instability waves along the SST front (Hansen and Paul, 1987). An inverse calculation using the hydrographic data compilation of Johnson, Sloyan, Kessler, and McTaggart (2002) suggested that about 10 Sv of equatorial water above the thermocline (including water upwelled from the upper layers of the EUC) may be transported northward by this process (Sloyan, Johnson, and Kessler, 2003). The whole question of interconnections among the EUC, SEC and Peru upwelling remains difficult to diagnose with presently available data.

Dynamically, it is not clear why there should be an EUC east of the Galapagos, because the usual understanding of the undercurrent suggests that it is driven by the zonal pressure gradient, which itself is due to the mean easterlies characteristic of the central Pacific. East of about 90°W, the zonal winds are very weak or in fact westerly (Fig. 1), so according to linear dynamics there should be no pressure gradient or EUC. (Note that the weak SEC in this region (Fig. 4) is consistent with the weak zonal winds). In fact the integrated zonal pressure gradient calculated from the AOML XBT data relative to 450m is reversed in the far east. This difficulty was recognized by Roden as early as 1962. A possible piece of the answer was proposed by Kessler, Johnson, and Moore (2003), who used an ocean GCM to show that eastward advection of cyclonic relative vorticity on the flanks of the EUC strengthened both the EUC and SEC branches well east of where an eastward undercurrent would be expected from linear dynamics, which is consistent with observations of these currents at 140°W and 110°W. While they had no data to compare with the far eastern Pacific mode results, this mechanism would act to produce an EUC in the far east even without a pressure gradient to drive it.

Fig. 1 shows the reversal of direction between thermocline-level currents (bottom panel) and surface currents (top panel) in the near-equatorial zone. At 100m, flow near 5°S and 5°N is eastward and spreading poleward; as has been noted in section 4.1 above, these are the Tsuchiya Jets, while the surface westward flow converging towards the equator is the SEC. The large vertical shear reflects the thick thermostad lying roughly between the 12°C and 14°C isotherms between 5°S and 5°N (Fig. 6, middle panel, and see Fiedler and Talley 2003 [this volume]), such that the lower isotherms bow down at the equator while the main thermocline bows up. Although geostrophic currents are not a reliable gauge of flow closer to the equator than about 3° latitude, the property tongues of Lukas (1986) suggested that the entire trans-equatorial region is also flowing eastward at thermocline level.

The role of the distinctive equatorial circulation in producing the vertical and horizontal exchanges that maintain the east Pacific cold tongue and the sharp SST front along about 2°N has been the subject of a great deal of research over 50 years (e.g., Cromwell 1953). Most of this work has concerned the central Pacific upwelling zone, but many of these processes are relevant to at least the part of the eastern equatorial Pacific west of the Galapagos, and the major points will be reviewed briefly here. The temperature balance is generally thought to be among equatorial upwelling and vertical mixing, meridional mixing across the SST front, and surface heating. Perhaps surprisingly, zonal advection seems to be less important. The continuous band of cool SST connecting the coast of Peru with the equatorial cold tongue might suggest advection of coastal water out along the equator. However, the southwestward surface currents west of Peru (Fig. 4), consistent with the Ekman drift implied by the mean winds (Fig. 1), do not support such a hypothesis. Some model experiments also contradict this idea. Kessler, Rothstein, and Chen (1998) put a wall extending 700km westward from the Ecuadorian coast at 4°S in an ocean GCM and found little difference in SST along the equator west of the Galapagos, either in the mean or the annual cycle. That implies that the cool SST is due to local upwelling in each location, which is plausible given the wind forcing, but cannot be considered confirmed. On the other hand, during El Niño events, zonal advection from the west can be a significant contribution to SST warming, at least in the central Pacific (Picaut, Ioualalen, Menkes, Delcroix, and McPhaden, 1996).

Numerous attempts to measure upwelling velocity have been based on the direct technique of finding the divergence of measured horizontal currents, e.g., Weisberg and Qiao (2000) used moored current meters, Johnson, McPhaden, and Firing (2001) used repeated shipboard ADCP sections, and Hansen and Paul (1987) used surface drifter currents. Indirect methods have been based on divergence of geostrophic and Ekman transports (Wyrtki, 1981; Bryden and Brady, 1985; Meinen, McPhaden, and Johnson, 2001). All these studies have estimated upwelling velocities on the order of a few m day^{-1}, with total vertical transport of about 50 Sv over the east-central Pacific, though the spatial pattern (whether broad and relatively slow or narrow and correspondingly fast) remains unknown. There is also little agreement on the depth to which upwelling reaches as it works against the stratification of the upper thermocline. Cross-isopycnal transport requires either heating from above (for example through penetrating radiation) or turbulent mixing, which is just beginning to be understood (see, e.g., Lien, Caldwell, Gregg, and Moum 1995). If not for the mixing and surface heating, upwelling would fill the equatorial upper layer with cool water. Despite the many uncertainties, there is no doubt that upwelling is an order(1) term in the heat and mass balance of the equatorial Pacific, both in the mean and in its seasonal and interannual variability. Since upwelling is so important but at the same time impossible to measure directly, attention has been given to consistency checks that compare the heat fluxes due to upwelling to other elements of the heat balance (Wang and McPhaden, 1999, 2000).

Mixing across the SST front north of the equator opposes the cooling due to upwelling. This mixing primarily takes the form of tropical instability waves (TIW), which are due to the shears between the eastward EUC and NECC and the westward SEC (Fig. 6), (Philander 1978; Willett and Leben 2003 [this volume]). It was first thought that the NECC-SEC shear was the key element, but some investigators now argue that the EUC-SEC shear is important as well (Yu, McCreary, and Proehl, 1995; Chelton, Schlax, Lyman, and Johnson, 2003), and this question remains controversial. The TIW distort the front into cusp-like shapes that are easily seen in satellite imagery (Legeckis, 1977; Chelton, Wentz, Gentemann, de Szoeke, and Schlax, 2000c; Chelton and Co-Authors, 2001), and result in a substantial equatorward heat transport by mixing across the front, as tongues of warm water move south and cool water move north. Hansen and Paul (1987) and Bryden and Brady (1989) estimated this heat flux as comparable to that due to upwelling, and other observational balances and model experiments agree.

The tendency to compensation between upwelling cooling and TIW warming occurs in the annual cycle as well. As winds strengthen in the second half of the year, the SEC and NECC both strengthen (see section 4.2 and Fig. 8), as does equatorial upwelling. These tendencies cool the equator and warm the region north of the SST front, increasing the SST gradient as well as strengthening the shear that drives the TIW and thereby produces stronger mixing. Therefore, during June-December, opposing SST tendencies are generated and the net effect of these ocean fluctuations on SST is smaller than might be assumed from the upwelling increase alone (Enfield, 1986; Kessler, Rothstein, and Chen, 1998). During El Niño events, by contrast, surface and local vertical fluxes are dominant (see next section).

Although the sun crosses the equator twice a year, in March and September, only an annual cycle is observed in SST, which is coldest in September (Mitchell and Wallace 1992; Fiedler and Talley 2003 [this volume]). Part of the reason for this is the cooling due to upwelling, which is maximum in boreal summer-fall, as noted above. Another aspect is the variations of the extensive stratus decks that cover the cool-water part of the eastern tropical Pacific, south of the SST front (Klein and Hartmann, 1993). The stratus clouds and cool SST comprise a positive feedback, since cool SST encourages the formation of stratus that blocks the usual September maximum insolation and therefore reinforces the SST anomalies. A similar situation prevails in the eastern Atlantic. Disentangling the diverse ocean-atmosphere influences on the eastern tropical oceans remains a central climate problem of the region that is just beginning to be attacked (Cronin, Bond, Fairall, Hare, McPhaden, and Weller, 2002).

4.4 El Niño variations

At the time of Wyrtki (1966), the warm SST anomalies that characterize El Niños along the coast of Peru were well known, but the extreme thermocline depth and eastward current anomalies along the equator were not realized until Wyrtki (1974a, 1975a) used island and coastal sea level time series to establish that the equatorial and coastal anomalies were part of the same event. The large anomalies produced by the El Niño of 1982-83 spurred a major effort that continues to the present to observe and understand the ENSO (El Niño-Southern Oscillation) phenomenon (see, e.g., McPhaden and Co-Authors 1998, Wallace and Co-Authors 1998 and Neelin and Co-Authors 1998 for reviews of this effort, and Wang and

Fiedler 2003 [this volume], for a review of recent theories). Today we know that El Niño events begin in the western Pacific and spread eastward, with the Peru coastal warming one of the final signatures. Although winds east of 110°W may remain close to normal throughout even a strong El Niño (Cronin and Kessler, 2002), drastic changes in the thermocline depth and currents still occur as the result of wind-forced variability carried to the east by equatorial Kelvin waves (Kessler and McPhaden, 1995b). In general, the westerly wind anomalies of the western and central Pacific flatten the thermocline across the basin, and at the same time greatly weaken the SEC (Fiedler, Chavez, Behringer, and Reilly, 1992; Kessler and McPhaden, 1995a; Lagerloef, Mitchum, Lukas, and Niiler, 1999; McPhaden, 1999; Wang and McPhaden, 2000; Bonjean and Lagerloef, 2002). SEC weakness occurs both because of weaker easterly winds and because the deepening equatorial thermocline depresses the thermocline bulge that produces westward geostrophic shear above the EUC (e.g., Fig. 6). The deeper thermocline allows local surface fluxes to warm the surface layer (Vialard, Menkes, Boulanger, Delecluse, Guilyardi, McPhaden, and Madec, 2001; Cronin and Kessler, 2002) as the source of cool upwelled water is cut off (though upwelling within the thick warm layer may still be occurring under the near-normal easterly trade winds). Eastward advection of warm west Pacific surface water appears to be a significant part of the El Niño warming in the west-central Pacific, but this does not penetrate far to the east (Picaut, Ioualalen, Menkes, Delcroix, and McPhaden, 1996; Bonjean, 2001). Although the warmest actual SSTs tend to occur in January-March of the year following the largest wind anomalies, warm SST and eastward zonal current anomalies have usually peaked in the previous September; thus El Niño appears in the east as a suppression of the cold phase of the annual cycle. It is not known why El Niño events are phase-locked to the annual cycle in this way (Wang and Fiedler 2003 [this volume]).

There is far less information to diagnose interannual variability in the eastern Pacific warm pool region, except for SST time series, for which instrumental records with sufficient spatial resolution to examine detailed features extend back to 1981 (Appendix, section C.2). The XBT data used in this study (Appendix, section A) were taken during the period 1979-96, but only along the ship track heading southwest from the Panama Canal is the sampling sufficient to produce interannual timeseries. Because of the paucity of subsurface data, Fiedler (2002) cited the results of a numerical model forced with observed winds to comment briefly on interannual variations of the Costa Rica Dome during 1980-2000 (he noted that it was weak or absent in El Niño years and was apparently stronger in the one La Niña year simulated by the model). Other data sets that have been used to study interannual variability in the tropical Pacific include island sea level records (Wyrtki, 1974b, 1979), but these provide less detail on the oceanic conditions in the east because the only long records are along the coast. For these reasons, studies of interannual variations in the northeastern Pacific has often relied on correlating basin-scale modes with local (coastal) time series (Baumgartner and Christensen, 1985). This is useful and informative but cannot speak to the variability of the currents or thermal structure away from the coast.

The lengthening time series from the Topex altimeter (Appendix, section C.3) gives a good picture of the evolution of the region since its launch in 1992, sampling at least the El Niño of 1997-98 and the change to La Niña conditions following that event, and we will use its spatial variability pattern to interpret interannual SST variability. Of course it must be kept in mind that the Topex variations are dominated by the very large signals of 1997-98.

A description of the spatial pattern of interannual variability of both the Topex sea level and SST is constructed by correlating interannual variability across the region with that at 0°, 95°W (Fig. 10), which is a good index for ENSO. For SST, correlations among the interannually-smoothed 20-year time series have about 15 degrees of freedom, which means that correlation values greater than about 0.5 are significantly different from zero at the 95% level, indicated by gray shading in Fig. 10. For sea level, there are fewer degrees of freedom, and correlations are significant above about 0.7. In either case, significant correlations indicate a close correspondence of interannual variations all along the equator and spreading poleward in a narrow region along the American coast. This is a classic signature of El Niño, in which deep thermocline anomalies (equivalently high sea level), accompanied by warm SSTs, propagate eastward along the equator as Kelvin waves, then poleward along the Americas. It is noted that a correlation map similar to that in Fig. 10 (top) was found from the Kaplan, Cane, Kushnir, Clement, Blumenthal, and Rajagopalan (1998) SST record that spans more than 100 years.

Several investigators have shown evidence of increased penetration of tropical water to the mouth of the Gulf of California associated with El Niño events (Baumgartner and Christensen (1985); Lavin, Palacios-Hernández, and Cabrera (2003)), consistent with Fig. 10, and sea level anomalies have been observed to propagate as far north as Alaska and as far south as Chile in response to the equatorial anomalies (Spillane, Enfield, and Allen, 1987).

It is noteworthy that the interannual SST correlation pattern is wider meridionally than the sea level pattern (Fig. 10), with high correlations extending to 8°-10° latitude, while for sea levels these extend only to about 6°-7° latitude. As mentioned above, within the eastern equatorial region, SST changes are primarily due to vertical processes associated with the deepening thermocline, however the correlation patterns suggest that on the flanks of the sea level anomaly the meridional height gradient drives eastward geostrophic currents that warm these extra-equatorial regions. This is consistent with previously-observed increases in the flow of the NECC, and corresponding weakness or reversal of the SEC at 5°-10°S, in the central Pacific during El Niños (Kessler and Taft, 1987; Kessler and McPhaden, 1995a), which are also associated with the increased depth of the near-equatorial thermocline at the height of the warm event. Indeed, when sea level anomalies from Topex during late 1997 are added to the seasonal XBT dynamic heights (Fig. 8), the surface height gradient across the NECC is approximately doubled, all the way up around the southern limb of the Costa Rica Dome, as is suggested by Fig. 10 (bottom). Further, since the sea level anomalies follow the coast narrowly, the Costa Rica Coastal Current is also greatly enhanced. However, this flow continues north along the coast rather than circling around the dome, thus should not be seen as a strengthening of the dome because it is due to thermocline deepening along the coast, not to shallow anomalies of the dome itself (the effect of El Niño sea level anomalies at the center of the dome is actually a slight thermocline deepening, consistent with Fiedler's (2002) findings from the numerical model results cited above).

5 Remaining questions

The central remaining physical oceanographic problem of the eastern tropical Pacific is the interconnection among the zonal currents of mid-Pacific as they impinge on, or depart from, the American coast. Wyrtki realized this early on, and drew a schematic (Fig. 5) that would not be drawn qualitatively much differently today; however that is mostly a statement that we are still unable to produce a quantitatively realistic picture.

We know, as Wyrtki did, that some of the EUC upwells into the SEC, while another part continues southeastward at thermocline depth to feed the Peru upwelling (Roden, 1962; Lukas, 1986) but are not much further along in putting numbers to this than was possible in 1966 (see Sloyan, Johnson, and Kessler 2003, for an estimate from an inverse calculation). In general the eastern origin of the SEC remains almost as shrouded in mystery as it was in Wyrtki's time. How much of the SEC comes from equatorial upwelling, how much from the NECC, and how much from the Peru coast? We can assume that this division varies both seasonally and interannually, but can barely speculate on the answers. A further puzzle concerns the very thick westward flows on both sides of the EUC, that appear to be deep extensions of the SEC branches (Fig. 6, middle panel). (Although these appear only weakly in the 90°W section (Fig. 6, bottom), they exist down to the 900m reference level at both 110°W and 90°W in this data set). Rowe, Firing, and Johnson (2000) hypothesized that these could be the westward limb of deep cyclonic gyres fed by the two Tsuchiya Jets. This has been difficult to confirm with the very sparse data coverage available.

Some of the most interesting questions, which bear directly on the role of this region in inter-basin and interocean exchanges, concern the vertical transports. As mentioned in section 4.3, water properties imply that EUC and much of the thermostad water below it originates in the southwest Pacific (see Fiedler and Talley 2003 [this volume]). The Indonesian Throughflow represents a transport of approximately 10 Sv; this amount of intermediate water entering the South Pacific leaves the North Pacific as surface water. Thus both the water mass must be transformed and its potential vorticity modified, and it must be upwelled, so that this transport can be effected. Both the water mass transformation and upwelling presumably occur in the eastern Pacific, as some combination of equatorial, Costa Rica Dome and Peru upwelling. This suggests that these distinctive features of the eastern tropical Pacific are in fact important elements of the global ocean circulation.

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Appendix: Data sources and preparation

A XBT data

The principal thermal data set used in this study is the historical XBT compilation of Donoso, Harris, and Enfield (1994), and will be referred to here as the "AOML XBT data". These profiles originated from earlier compilations produced principally by the National Ocean Data Center, the Joint Environmental Data Center at Scripps, and the Global Subsurface Data Center in Brest, France. A total of 36185 profiles were available within 15°S-25°N, and from 120°W to the American coast; as is typical of volunteer observing ship data, the profiles were primarily organized along the major shipping routes. The main concentration of profiles were taken during 1979 through 1992, with smaller numbers after that date. Donoso, Harris, and Enfield (1994) describe the objective and subjective quality control procedures used to flag questionable observations and produce a research-quality data set. Only data points they found to be acceptable were used in this study.

The irregularly-distributed profiles were mapped to a regular 1° latitude by 1° longitude by 10m by climatological-month grid in two steps. First, the temperature data in each individual profile were linearly interpolated to 10m depth intervals from the surface to 450m (on average, each profile had about 41 samples in the upper 450m). Second, in a separate calculation at each 10m level z_0, the irregularly-spaced temperatures were mapped onto the (x, y, z_0, t) grid using a Gaussian-weighted three-dimensional mapping (Kessler and McCreary, 1993), with scales of 2° longitude, 1° latitude and 2 months. The region of influence of each observation was limited to where its weighting was greater than e^{-4}, that is, to twice the scale distance from the center of each gridbox. Only the month was considered in the gridding, not the year, to produce a climatological average annual cycle.

Although some earlier XBT-based climatologies of the entire tropical Pacific used coarser grid spacings [e.g., Kessler 1990, who mapped onto a 5° longitude by 2° latitude grid], the short spatial scales of the phenomena of interest in this region required higher resolution, and the profile distribution in the new compilation was dense enough to justify a 1° by 1° spacing. Each 1° by 1° gridbox contained, on average, more than 24 profiles during the year, spread over an average of 7.4 months. The gridded temperature climatology was checked by comparison against other sources, including the Reynolds SST (Reynolds and Smith, 1994) and time series from the TAO mooring array (McPhaden and Co-Authors, 1998).

Dynamic height relative to 300m or 450m reference levels was found from the gridded temperatures via a mean T-S relation constructed from the Levitus, Burgett, and Boyer (1994) atlas, and geostrophic currents were found from centered differences of the dynamic heights.

An additional meridional section at 125°W, shown in Fig. 6, was constructed from the XBT data compilation of Kessler (1990). These data partly overlap the AOML XBT compilation, but extend further west. Quality control and gridding is described by Kessler (1990); of note is that the lesser data density required a coarser gridding as mentioned above. Along the 125°W section shown, 50-150 profiles were found in each 2° latitude by 5° longitude box, roughly half as dense than the AOML data.

B Scatterometer wind data

Winds are used in this study to estimate Ekman pumping and the Sverdrup balance. Since the winds in this region are known to have small spatial and temporal scales (Chelton, Freilich, and Esbensen, 2000a), especially in the case of the mountain-gap wind jets that will be seen to have great importance here, and because these calculations require spatial derivatives of the winds, excellent sampling is a necessity. This points to the utility of satellite scatterometer winds that can resolve the wind jets better than any in situ or present-generation reanalysis wind product (Schlax, Chelton, and Freilich, 2001). The Quikscat satellite carrying a Seawinds scatterometer (referred to here as Quikscat) was launched in June 1999 and its operational products became available in late July 1999. The satellite circles the earth in a 4-day repeat, sun-synchronous orbit with a spacing of 6.3° at the equator, sampling a 1600 km-wide swath centered on its ground track.. The first three years of Quikscat wind stresses were obtained on a 1° by 1° grid from the web site of Center for Ocean-Atmosphere Prediction Studies (COAPS) at Florida State University (FSU), and used to represent the mean winds.

The European Research Satellite (ERS) was launched in July 1991, and its nearly identical follow-on mission was launched in April 1995 and remains in operation. ERS data are processed onto a 1° by 1° monthly grid by the French data processing facility Centre ERS d'Archivage et de Traitement (CERSAT), and were obtained by ftp from the CERSAT website (http://www.ifremer.fr/cersat). The ERS sampling is not as good as Quikscat, with a 500 km- wide swath and 35-day repeat cycle. However, the more than ten year record is more suitable for estimating the average annual cycle, and the ERS winds are used here to force a Rossby wave model (section 4.2.1). Extensive comparison has been made between the two scatterometer products, and also with *in situ* wind products (the widely-used Florida State University product, which spans 40 years (Stricherz, O'Brien, and Legler, 1992)). All these winds show quite a consistent picture in the eastern tropical Pacific; the differences are essentially that Quikscat gives greater spatial detail than ERS, which in turn gives greater detail than FSU. Given this, we have chosen to use Quikscat to represent the mean winds, since the spatial detail is valuable to show the effects of the Central American wind jets, but to use ERS for the average annual cycle used to force the Rossby wave model since its longer record suggests a better resolution of the time variability.

C Supplemental data sets

A few additional data sets are used for particular purposes. These have been extensively described in the literature and are briefly summarized here.

C.1 Directly measured subsurface velocity

Directly measured velocities are shown in Fig. 6 at 110°W and 125°W, from the sections constructed by Johnson, Sloyan, Kessler, and McTaggart (2002). These data were taken from an ADCP (Acoustic Doppler Current Profiler) mounted on the ship servicing the TAO moorings along those longitudes. Twelve sections were made along 110°W and 14 along 125°W, within 8°S-8°N, during 1991-2001. Data processing is described in Johnson, Sloyan, Kessler, and McTaggart (2002); an objective mapping along isopycnals produced a mean, annual cycle, and ENSO variability estimate. Here, just the mean zonal velocity along two sections is shown (Fig. 6, top two panels) to supplement the geostrophic velocity estimates from XBT data.

The TAO mooring at 0°, 110°W has been instrumented with current meters of various types since the early 1980s. Until 1997, the surface mooring included Vector Averaging Current Meters (VACMs), typically mounted at eight depths from 10m to 250m. Since 1991, velocities have been measured using Acoustic Doppler Current Profilers (ADCPs), which provide much finer vertical resolution (typically 8m; McPhaden, Milburn, Nakamura, and Shepherd 1991). At first the ADCPs were mounted on the surface mooring to look downward; since 1995 they have been mounted on subsurface platforms (300m) looking upward, which alleviates damage from fishing operations around the surface mooring.

C.2 Reynolds SST

A gridded sea surface temperature (SST) product based on satellite AVHRR sampling ground- truthed with in situ data is produced by the National Centers for Environmental Prediction on a weekly, 1° by 1° grid for the period Oct. 1981 through the present. Reynolds and Smith (1994) describe the data processing and quality control procedures. This data set is commonly known as the “Reynolds SST”, and we will use that nomenclature here.

C.3 Topex altimetric sea level anomalies

The TOPEX/POSEIDON altimeter has measured sea surface height (SSH) nearly continuously since October 1992. Its orbit repeats a diamond formed by the overlapping patterns of ascending and descending tracks every 9.9 days; in the tropics the diamond spans about 2.8° latitude and 7.8° longitude. Fu and Co-Authors (1994) describe the technical characteristics and measurement accuracy of the instrument.

C.4 Drifter surface currents

Surface currents estimated from drifters are used as a check on the geostrophic currents derived from XBT data. Surface Velocity Program (Niiler, Sybrandy, Bi, Poulain, and Bitterman, 1995) drifter data were obtained from the Atlantic Oceanographic and Meteorological Laboratory as 6-hourly kriged positions and velocities (Hansen and Poulain, 1996). 1045 drifter tracks were found in the study region, however the distribution is not ideal since the drifter population falls off sharply closer than about 1000 km to the Central American coast. The region west of Cabo Corrientes and south of Baja California is especially poorly sampled. Nevertheless, the 6- hourly velocities were mapped to a 1° by 1° by monthly climatological grid by the same method and with the same scales as was done for the XBT temperatures described in section A.

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